A subduction origin for komatiites and cratonic lithospheric mantle

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ABSTRACT

We present a model in which the generation of komatiites in Archaean subduction zones produced depleted mantle residues that eventually formed the highly depleted portions of the Kaapvaal lithospheric mantle. The envisioned melting process is similar to that which has formed boninites in Phanerozoic subduction zones such as the Izu-Bonin-Mariana arc. The primary differences between the Archaean and Phanerozoic melting regimes are higher mean melting temperatures (1450 versus 1350 °C) and higher mean melting pressures (2.5 versus 1.5 GPa) for the komatiites. The komatiites from the Komati Formation in the Barberton greenstone belt are mafic enough to have produced the depletion seen in most Kaapvaal granular peridotite xenoliths. However, the most highly depleted Kaapvaal xenoliths require an even more Mg-rich magma than the Komati komatiites (Kk). Samples of boninite mantle residues from the fore-arc of the Marianas subduction zone are nearly as depleted as the Kaapvaal cratonic mantle, indicating that buoyant, craton-like mantle is being produced today. We speculate that production rates of cratonic mantle were greater in the Archaean due to the greater depth of melting for komatiites (relative to boninites) and greater worldwide arc length. The high production rates and high buoyancy of the komatiite mantle residues gave rise to the rapid growth and stabilization of the Kaapvaal craton in the Archaean.

Introduction

The lithospheric mantle that underlies the Kaapvaal craton has chemical and physical properties distinct from those of the convecting mantle (Boyd, 1989). The composition of mantle xenoliths, brought to the surface by kimberlites, suggest that the cratonic mantle was extensively melted at some point in its history. Dating of mineral inclusions in diamonds suggests the depletion event occurred in the Archaean (3.3 to 3.2 Ga; Richardson *et al.*, 1984; Shirey *et al.*, 2004). Even though the high degree of melting recorded by the samples should have removed garnet (gt) from the mantle residue, garnet is present in many of the xenolith samples. The garnets are spatially associated with orthopyroxene (opx) and clinopyroxene (cpx), and it is now thought that the garnets were produced by exsolution from opx (Cox *et al.*, 1987). Reconstructions of the unexsolved opx compositions suggest that the melting event occurred at low pressures (1.5 to 2.5 GPa) and high temperatures (>1400°C), after which the residue was cooled and taken to higher pressures (Saltzer *et al.*, 2001). Subsequent to the melting event, metasomatism of various types has affected the cratonic mantle, changing the modal proportions of minerals and enriching the mantle in minor and trace elements (Nixon and Davies, 1987). In particular, many of the Kaapvaal xenoliths have high proportions of opx, higher than would be expected based on their degree of depletion (Boyd, 1989). A variety of hypotheses have been forwarded to explain the observed opx enrichment, though there is no consensus about which one is most likely. The ideas include metamorphic differentiation (Boyd, 1989), melting at high pressures (Kinzler and Grove, 1999), addition of SiO_2 by subduction fluid (Kesson and Ringwood, 1989b) and melt-wallrock reaction (Kelemen *et al.*, 1992; 1998).

Physically, both seismic and petrologic studies indicate that the Kaapvaal lithosphere is extremely thick (250 to 300 km; James *et al.*, 2001; Stankiewicz *et al.*, 2002). Re-Os dating indicates that the lithosphere has been stable since the Archaean (Pearson *et al.*, 1995), and so the Kaapvaal lithospheric 'keel' appears to have been resistant to plate tectonic forces over most of the Earth's history. Seismic tomography (James *et al.*, 2001) and geothermometry on xenoliths (Finnerty and Boyd, 1987) both indicate that the lithosphere is cold, relative to the convecting mantle. In general, the cold temperatures should cause the mantle to be dense and therefore buoyantly unstable. It should sink. But due to the high degree of depletion of the cratonic mantle, the concentrations of FeO and Al_2O_3 are low (relative to the asthenophere), reducing the bulk density of the mantle. Interestingly, the Kaapvaal crust does not record large vertical motions over its ~3.6 Ga history, leading researchers to conclude that the positive buoyancy from the chemical depletion offsets the negative buoyancy from the cold thermal state of the lithospheric mantle (Jordan, 1981); the so-called 'isopycnic hypothesis'.

Table 1. Major element composition of komatiite chill margins from the Komati Formation

Sample	**K6-1KC**	**K6-1KCA**	**K6-1KC2**	**K6-1KCB**	**K6-1KAB**	**K6-1KAA**	**K6-1KBA**	**K6-1KBB**
Horizon	**1**	**1**	**1**	**1**	**1**	**1**	**1**	**1**
SiO_2	44.8	44.61	45.5	45.51	45.78	45.41	44.97	45.38
TiO_2	0.20	0.19	0.20	0.20	0.21	0.23	0.23	0.24
Al_2O_3	3.31	3.43	3.46	3.37	3.48	3.86	3.92	4.07
Cr_2O_3	0.45	0.46	0.47	0.48	0.46	0.47	0.47	0.48
Fe_2O_3	12.74	14.09	12.87	13.11	13.55	13.67	14.63	14.05
MnO	0.19	0.21	0.20	0.20	0.21	0.22	0.23	0.21
MgO	32.71	31.54	31.53	31.12	29.62	29.51	29.41	28.69
CaO	5.10	5.08	5.37	5.56	6.11	6.00	5.59	6.26
Na_2O	0.20	0.13	0.14	0.17	0.30	0.34	0.28	0.34
K_2O	0.02	0.02	0.02	0.01	0.03	0.04	0.03	0.04
P_2O_5	0.02	0.02	0.02	0.02	0.02	0.03	0.03	0.03
NiO	0.26	0.23	0.24	0.23	0.23	0.22	0.21	0.21
Total	100	100	100	100	100	100	100	100
LOI	4.53	5.03	5.26	4.85	4.47	4.77	5.92	5.18
Sample	**K4-1LWM**	**K4-1BA**	**K4-1B**	**K5-18A**	**K5-18AD**	**K4-47B**	**K4-1A2B**	**K1-19**
Horizon	**2**	**2**	**2**	**2**	**2**	**2**	**2**	**2**
SiO_2	46.84	46.21	47.29	46.88	47.00	46.87	47.14	47.46
TiO_2	0.31	0.38	0.37	0.36	0.36	0.34	0.37	0.39
Al_2O_3	3.07	4.37	3.66	4.04	3.80	3.47	3.85	3.96
Cr_2O_3	0.31	0.36	0.34	0.35	0.34	0.36	0.35	0.35
Fe_2O_3	12.00	12.56	12.42	12.44	12.59	13.06	12.62	12.72
MnO	0.17	0.18	0.17	0.15	0.17	0.19	0.20	0.20
MgO	31.6	29.9	29.33	29.12	28.84	28.39	27.38	27.32
CaO	5.40	5.71	6.09	6.20	6.48	6.51	7.61	7.03
Na_2O	0.03	0.07	0.09	0.16	0.15	0.53	0.26	0.31
K_2O	0.02	0.01	0.02	0.08	0.05	0.05	0.03	0.04
P_2O_5	0.02	0.03	0.03	0.03	0.03	0.03	0.03	0.03
NiO	0.23	0.21	0.20	0.19	0.19	0.19	0.17	0.18
Total	100	100	100	100	100	100	100	100
LOI	7.69	6.54	6.48	6.31	6.83	4.51	5.93	6.13
Sample	**K4-1A2E**	**K4-1E1**	**K4-1A**	**K4-1DE**	**K4-1DK**	**K4-1DH**	**K4-1E2**	**K4-1DC**
Horizon	**2**	**2**	**2**	**2**	**2**	**2**	**2**	**2**
SiO_2	47.45	47.96	47.38	47.91	48.14	49.3	48.79	49.35
TiO_2	0.37	0.37	0.38	0.40	0.39	0.38	0.37	0.40
Al_2O_3	3.77	3.51	3.94	3.98	3.92	3.72	3.21	3.69
Cr_2O_3	0.35	0.36	0.37	0.32	0.33	0.33	0.36	0.33
Fe_2O_3	12.56	12.48	13.01	12.98	12.59	11.09	12.45	11.34
MnO	0.21	0.18	0.20	0.17	0.17	0.17	0.17	0.17
MgO	26.53	26.12	24.94	24.63	24.6	24.36	24.31	24.19
CaO	8.33	8.59	9.39	9.06	9.30	10.08	9.83	10.00
Na_2O	0.19	0.18	0.17	0.29	0.30	0.29	0.23	0.26
K_2O	0.03	0.02	0.02	0.02	0.02	0.04	0.02	0.03
P_2O_5	0.03	0.03	0.03	0.04	0.04	0.03	0.04	0.05
NiO	0.17	0.22	0.17	0.21	0.20	0.20	0.22	0.20
Total	100	100	100	100	100	100	100	100
LOI	5.66	3.95	4.74	3.08	3.02	3.37	2.84	3.09

Table 1. Major element composition of komatiite chill margins from the Komati Formation (continued)

Sample	**B96-13A**	**B94-6561**	**K4-1A2A**	**K4-1UPC**	**K4-1LWC**	**K5-22C**	**K5-22F**	**K5-22E**
Horizon	**2**	**2**	**2**	**2**	**2**	**3**	**3**	**3**
SiO_2	49.71	47.99	47.86	48.02	48.72	49.6	49.61	48.76
TiO_2	0.40	0.39	0.39	0.43	0.36	0.36	0.36	0.43
Al_2O_3	3.26	3.68	3.68	4.14	3.30	3.68	3.67	4.32
Cr_2O_3	0.33	0.37	0.38	0.38	0.35	0.35	0.35	0.39
Fe_2O_3	12.33	13.12	13.39	13.51	12.48	12.41	12.55	12.76
MnO	0.18	0.19	0.20	0.22	0.18	0.15	0.15	0.16
MgO	24.15	24.02	23.88	23.47	23.36	28.22	27.33	27.19
CaO	9.30	9.96	10.00	9.42	10.62	4.86	5.62	5.67
Na_2O	0.12	0.05	0.01	0.19	0.28	0.00	0.00	0.00
K_2O	0.02	0.02	0.02	0.02	0.08	0.13	0.10	0.13
P_2O_5	0.04	0.03	0.03	0.04	0.05	0.03	0.04	0.03
NiO	0.16	0.17	0.17	0.16	0.22	0.21	0.21	0.16
Total	100	100	100	100	100	100	100	100
LOI	4.42	4.55	4.31	3.90	3.48	5.64	5.49	5.75

Sample	**K5-22B**	**B96-14**	**K5-10(2)**	**K15-6e**	**K13-3DB**	**K13-3DA**
Horizon	**3**	**3 or 4**	**4**	**5**	**5**	**5**
SiO_2	52.07	47.59	46.9	44.86	47.38	46.96
TiO_2	0.33	0.48	0.40	0.43	0.38	0.37
Al_2O_3	3.43	5.25	4.07	4.50	3.15	3.24
Cr_2O_3	0.34	0.37	0.36	0.37	0.34	0.37
Fe_2O_3	11.17	13.32	12.47	13.82	12.97	13.11
MnO	0.13	0.20	0.20	0.17	0.18	0.18
MgO	26.71	24.2	26.21	28.66	28.38	27.75
CaO	5.44	8.08	8.99	6.75	6.95	7.76
Na_2O	0.00	0.27	0.05	0.01	0.00	0.00
K_2O	0.13	0.09	0.16	0.20	0.04	0.04
P_2O_5	0.03	0.04	0.03	0.04	0.03	0.03
NiO	0.21	0.11	0.15	0.18	0.20	0.20
Total	100	100	100	100	100	100
LOI	4.73	4.04	5.43	7.03	6.37	6.7

All values in weight percent and analyses normalized to 100. LOI is loss on ignition and is the sum of the volatile content of the rock. LOI gives a general idea of the degree of metamorphism a sample has undergone. All samples were analyzed by XRF at the University of Durban by Allan Wilson. Horizon refers to the five horizons of olivine spinifex komatiites in the lower Komati Formation and are numbered starting at the lowest stratigraphic level.

Ideas regarding the formation and subsequent evolution of the Kaapvaal lithosphere can generally be broken into two categories: 1) formation above a mantle plume (*e.g.* Boyd, 1989; Herzberg, 1993), 2) formation by successive underthrusting of oceanic crust in convergent margins (*e.g.* Abbott, 1991; de Wit *et al.*, 1998; Helmstaedt and Schulze, 1989). The former models generally argue for a hot Archaean dominated by plume activity, with the residues of the super-hot plumes forming the cratonic mantle. The second class of models argues for plate tectonic-like activity in the Archaean. Proponents of this view note that the compositions of peridotite and eclogite xenoliths are consistent with an origin as altered oceanic lithosphere (Schulze, 1986; Shirey *et al.*, 2001). While the topic is highly controversial, we believe the weight of evidence favors the subduction origin (de Wit *et al.*, 1998).

All of the models implicate komatiites as the magmas that caused the high degree of depletion of the cratonic mantle. Komatiites are highly mafic magmas that were produced almost exclusively in Archaean and Proterozoic times. These magmas have higher liquidus temperatures than any modern magma and have been used as primary evidence that the Earth's mantle has cooled since the Archaean (Herzberg, 1995). Three observations suggest a link between komatiites and the cratonic mantle: 1) The extent of depletion recorded by the Kaapvaal xenoliths is consistent with extraction of a komatiite-like magma. No other magma known is mafic enough to have produced the cratonic mantle, 2) Komatiites are found almost exclusively in the crust of cratons, above the depleted lithospheric mantle which they are proposed to have formed. Though horizontal crustal motions could have juxtaposed the two, the repeated spatial association of komatiites and cratonic mantle suggests a genetic relationship, 3) The ages of the oldest komatiites on the Kaapvaal craton (~3.49 Ga; Lopez-Martinez *et al.*, 1992) and the Re-Os ages for

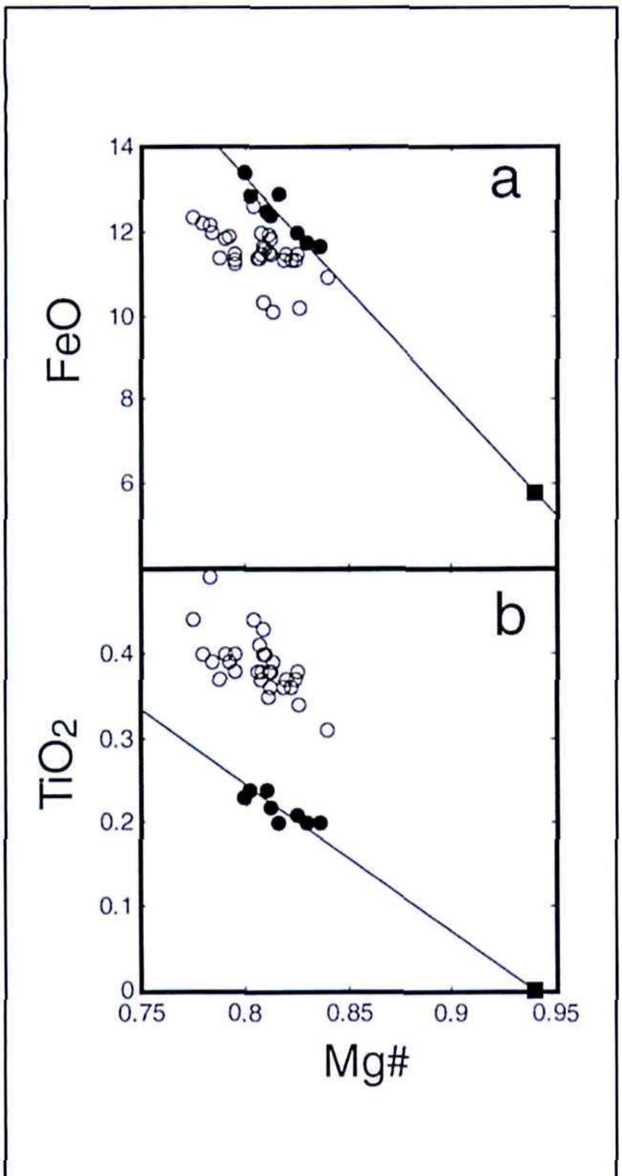

Figure 1. Compositions of komatiite chill margins from the Komati Formation. (Table 1). **(a)** FeO content versus Mg# (=MgO/MgO+FeO). **(b)** TiO_2 content versus Mg#. The stratigraphically lowest horizon of olivine spinifex komatiites (h1, filled circles) has distinctly lower TiO_2 contents and slightly higher Mg#s than the other four olivine spinifex horizons (h2-h5, open circles). The compositional trends in the h1 samples are consistent with fractionation of the most Fo rich olivine found in the Komati Formation (filled square). The unfractionated h1 magmas are estimated to have had 11.0 to 11.7wt.% FeO, while the h2-h5 magmas had 11.0 to 11.4wt.%.

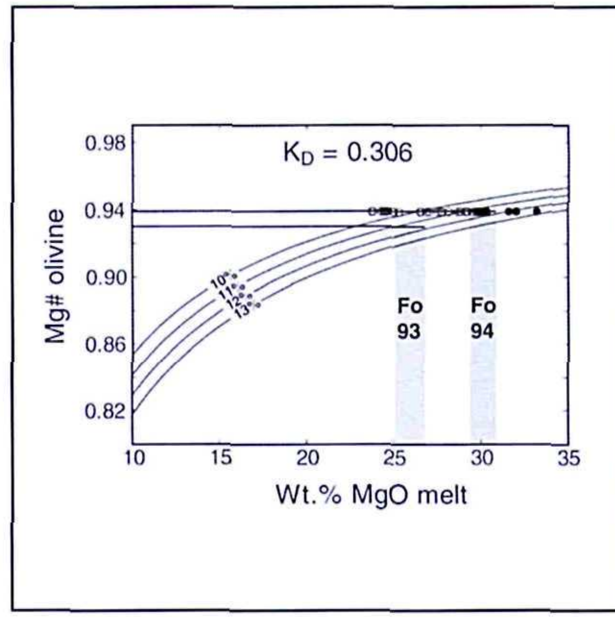

Figure 2. Mg# of olivine in komatiites versus the MgO content of *the melt it is in equilibrium with, using an Fe-Mg K_D of 0.306.* Curves show calculation for varying FeO contents (labeled). The upper horizontal line is the highest Mg# ol (0.939) in the h1 komatiite samples. The lower horizontal line is the lower limit of average olivine core compositions (Mg#≈0.930) for all of komatiite samples. Vertical lines show the upper and lower limits on the MgO contents of the melts in equilibrium with the corresponding olivine. MgO contents of Kk chill margins from h1 (filled circles) and h2-5 (open circles) are plotted along the Mg# = 0.939 line for comparison. The most Fo rich olivine could have been in equilibrium with melts with up to 31 wt.% MgO.

Kaapvaal mantle xenoliths (3.3-3.5 Ga; Pearson *et al.*, 1995) indicate that they were formed at approximately the same time.

Komatiites are widely thought to have been produced in exceptionally hot, deep-seated plumes (Arndt *et al.*, 1998; Herzberg and Ohtani, 1988). Their anhydrous liquidi are >1600°C and suggest mantle temperatures 200 to 400°C hotter than in modern plumes (Herzberg and Ohtani, 1988). Such high geotherms intersect the mantle solidus at greater depths than modern plumes (>10 GPa; Walter, 1998) and adiabatically melt over much larger intervals, producing high extents of melting (30 to 60%). The origin of komatiites in plumes is the cornerstone of the hypothesis that the cratonic mantle also formed in plumes.

The plume model for the origin of komatiites however, is not readily compatible with a subduction origin for the cratonic mantle. If the cratonic lithosphere represents subducted Archaean oceanic mantle, then the overlying komatiites (the oceanic crust) would have been subducted also. In that case, the komatiites preserved in greenstone belts would not be directly related to the cratonic mantle, but would have to be unusual oceanic crust (maybe back-arc) that was obducted onto the continents. In addition, the thermal buoyancy of a plume big enough to form the cratonic mantle would cause a large amount of uplift and consequent erosion. As the plume waned, the lithosphere would subside, creating a basin 10 to 15km deep (Jordan, 1981), yet no evidence of such a large sedimentary basin or thick crust (Nguuri *et al.*, 2001), has been found in the Kaapvaal.

The difficulty in reconciling a plume origin for komatiites and a subduction origin for the cratonic

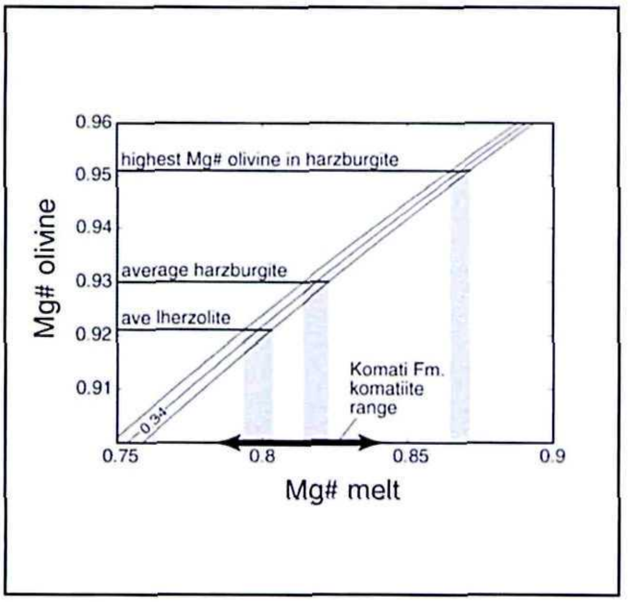

Figure 3. Mg# of olivine in the Kaapvaal mantle versus the Mg# of komatiite magmas for K_D = 0.340±0.010 (curves). The average Mg# of olivines in granular peridotites and harzburgites is 0.921 (lower horizontal line) and 0.930 (middle horizontal line), respectively, (Danchin *et al.*, 1979). The Kk (Mg# =0.78-0.83) could have produced these degrees of depletion. But the highest degrees of depletion in the Kaapvaal xenoliths (Mg# = 0.95) requires melts more mafic than the Kk.

mantle was recognized by Kesson and Ringwood (1989a). In their words, the Archaean lithosphere, "... evolved via magmatic processes characterized by higher degrees of partial melting and the extensive generation of komatiitic magmas [in plumes]... Alternatively, (and probably in addition) the diamond protolith [cratonic mantle] may have been generated beneath the Archaean equivalents of island-arc systems." This belies the observation that it is logically desirable to have the cratonic mantle be produced by subduction-related magmatic events, but that the only candidate magmas are apparently produced by plumes.

A subduction origin for komatiites

As part of the Kaapvaal Project (Anonymous, 2001), we have re-examined the geology, petrology and geochemistry of the komatiites and related rocks of the Komati and Hooggenoeg Formations, the type-localities for komatiites (Viljoen and Viljoen, 1969). Our experimental study of a least-altered komatiite composition indicates that the preserved igneous minerals in the Komati samples can only have been produced if the komatiite magmas contained >3 wt.% H_2O (Parman *et al.*, 1997). This is significantly higher than any modern plume magma (<0.5 wt.% H_2O). Initially, we proposed that the high H_2O contents indicated that either 1) the Komati komatiite (Kk) magmas were produced in an Archaean subduction zone or 2) that at least some Archaean plumes were hydrous and that the water was left-over from the accretion of the Earth. Subsequently, geochemical analyses of the komatiites and inter-layered komatiitic basalts and low-Ti tholeiites have indicated a link between the komatiitic magmas and modern boninites (Parman *et al.*, 2001).

Boninites are the products of high-degree, hydrous melting in subduction zones (Crawford *et al.*, 1989). They are perhaps the most mafic modern magmas, with estimated MgO contents up to 20 wt.% (Sobolev and Danyushevsky, 1994). Experimental studies agree that they were produced by hydrous melting at relatively low pressures (1 to 2 GPa; Falloon and Danyushevsky, 2000). Their extremely depleted trace and minor element compositions argue strongly that they were melts of a previously depleted mantle (Hickey and Frey, 1982), but that many were enriched in large ion lithophile elements (LILE) by the hydrous fluid that instigated the mantle melting event. Initial H_2O contents for boninites are estimated to be 2 to 4 wt.% (Ohnenstetter and Brown, 1996), similar to our estimates for the Kk (>3 wt.%).

Boninites are not produced in typical subduction environments. Most of the autochthonous boninites are found in the fore-arcs of western Pacific convergent margins. In the Marianas, the boninites were produced during the initiation of subduction, before the fore-arc could be extensively cooled by the downgoing lithosphere (Stern and Bloomer, 1992). In some cases, boninites are produced in back-arcs, where the rift zone propagates into hydrated mantle (Falloon and Crawford, 1991). While the time over which boninites were produced in the western Pacific was short, the magma production rates were extremely high (Stern and Bloomer, 1992). Boninite crust and mantle appears to be resistant to subduction as it is preserved within continental crust (as ophiolites) throughout the world (Crawford *et al.*, 1989).

Chemical similarities of komatiites and boninites

The major element compositions of boninites coincide with the compositions of Komati komatiitic basalts for all elements (Parman *et al.*, 2001). In general, basaltic komatiites are petrographically similar to boninites (Cameron *et al.*, 1979). Key trace element characteristics of boninites, such as low Ti/Zr, and Zr-Hf enrichments (relative to REE) are found in both the Komati komatiitic basalts and in the komatiites themselves (Parman *et al.*, 2001). Rare Barberton komatiitic basalts have the U-shaped REE patterns characteristic of many boninites (Jahn *et al.*, 1979). In addition, boninites are typically interbedded with low-Ti (back-arc) tholeiites (Bedard, 1999; Crawford and Keays, 1987). Nearly identical tholeiites are found interlayered with the komatiitic rocks of the Komati and Hooggenoeg Formations. Worldwide, Archaean boninitic lavas have been recognized in Archaean greenstone belts, sometimes inter-layered with komatiites (Kerrich *et al.*, 1998; Poidevin, 1994).

The above observations were used by Parman *et al.* (2001) to propose that the Komati Formation komatiites

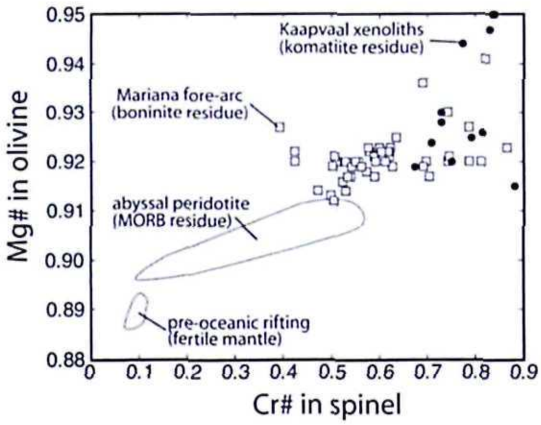

Figure 4. Mg# of olivines versus the Cr# (=Cr/(Cr+Al)) of spinels in mantle samples. Abyssal peridotites (large gray field) is depleted relative to fertile mantle (small gray field; Bonatti and Michael, 1989). Residues of boninite melting (squares) are more depleted still, with a large amount of overlap with the Kaapvaal xenoliths (circles).

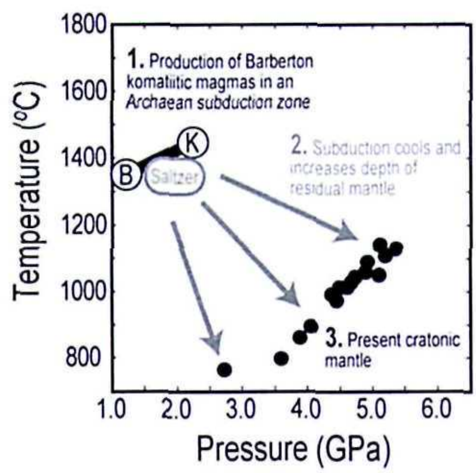

Figure 5. Pressure-temperature evolution of the Kaapvaal lithospheric mantle. Komatiite (**K**) and boninite (**B**) melting conditions (Parman, 2001) are very similar to mantle conditions for Kaapvaal xenoliths (grey oval; Saltzer *et al.*, 2001) estimated by reconstructing opx compositions from exsolved gt and cpx. Arrows indicate the cooling and burial of the mantle residues of melting, ending at the inferred conditions for the current cratonic mantle (filled circles). See text for a more complete description.

were produced *in the Archaean equivalents of* subduction zones, by a similar melting process that produced boninites. High pressure, hydrous phase equilibria experiments indicate that the difference between boninite melting conditions and komatiite melting conditions is ~100°C, and so the Archaean mantle was slightly hotter than it is today. This is consistent with numerical models of the thermal evolution of the Earth (Davies, 1995) and with empirical estimates of Archaean mantle temperatures (Abbott *et al.*, 1994).

In this contribution, we examine the implications that a subduction origin for komatiites has for the formation of the Kaapvaal lithosphere. We show that previous models of the formation of the cratonic mantle by subduction processes are made much less complicated if komatiites are subduction magmas. In addition, we show that modern boninites are producing mantle that is nearly as depleted (and therefore as buoyant) as the Archaean cratonic mantle. This raises the question, "Why are there no young cratonic keels?"

Composition of Komati Formation komatiites

Before discussing the subduction hypothesis, we use new chemical data on the Kk (Table 1) to evaluate the concept that these komatiites are mafic enough to have produced the depleted cratonic mantle. The MgO content of komatiite magma is a key value, as it is directly proportional to the liquidus temperature, and therefore to estimates of the melting conditions and to the overall extent of melting recorded by the komatiites. All komatiites, including those from the Komati Formation, are heavily metamorphosed. Serpentine and chlorite are the main alteration products, with the most MgO rich samples having the most serpentine and the highest metamorphic H_2O contents (de Wit, 1987; Parman *et al.*, 1997). This suggests that the MgO contents of the samples could have been increased during metamorphism. One way to constrain the unmetamorphosed composition of the magmas is to test whether the preserved igneous olivine in the samples would have been in equilibrium with a melt that had the bulk composition of the sample.

Fe-Mg exchange between olivine (ol) and melt can be used to test for equilibrium between olivine and melt using the distribution coefficient K_D $=(FeO/MgO)^{ol}/(FeO/MgO)^{melt}$ (Ulmer, 1989; Walter, 1998). For mafic melts at 1 atm, the value is ~0.300. Fe-Mg K_D is influenced by oxygen fugacity. Estimates of the oxygen fugacity of the Kk magmas have been estimated from the composition of chromian spinels and indicate an fO_2 of ~QFM (Canil, 1997). Our experimentally determined Fe-Mg K_D value for a Kk composition at QFM is 0.306 (Parman *et al.*, 1997).

To determine the MgO content of the komatiite liquid using the Fe-Mg K_D, one must first know the FeO content of the sample. The bulk FeO contents of the chill margins of Kk units vary from 10 to 13.5wt.% (Figure 1a). Fractionation of olivine should produce a

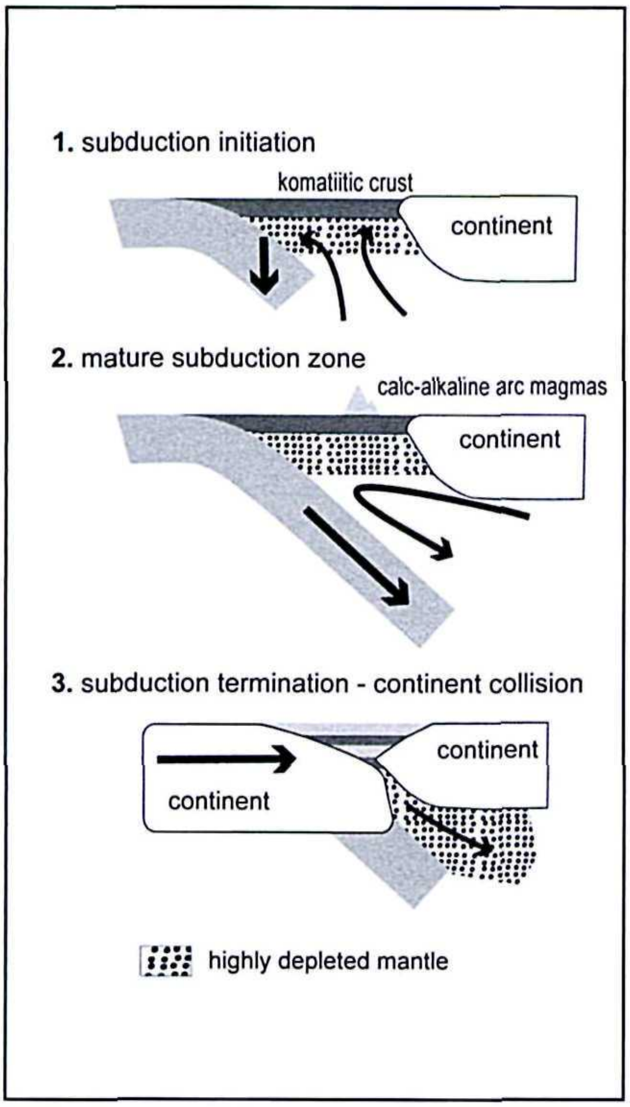

Figure 6. Cartoon of proposed process for the generation of komatiites and cratonic mantle by subduction-related melting. **(1)** Boninite-like melting event produces komatiite, komatiitic basaltic magmas and low-Ti tholeiites (dark shading) and corresponding heavily depleted mantle (stippled region). **(2)** Mature subduction cools and hydrates residual mantle which is stuck in the fore-arc. **(3)** Komatiitic crust is obducted onto continents during continent collision at the end of subduction. The depleted mantle is thickened and incorporated into the continental lithospheric mantle. It is possible or even likely that some portion of the komatiite residual mantle would be obducted along with the corresponding volcanic section, forming a more complete ophiolite sequence as proposed by de Wit *et al.* (1987).

strong increase in the FeO contents of the melts, but grouping all of the data together, no trend is apparent. In the Komati Formation, the olivine-spinifex komatiites occur as five horizons within the Lower Komati Formation (horizons h1-h5; Dann, 2000). The lowest horizon stratigraphically (h1) is compositionally distinct from the overlying four horizons, with lower TiO_2 contents and higher Mg#'s (Figure 1b; Mg#=MgO/(MgO+FeO)). The FeO contents of the chill margins of h1 do form a fractionation trend (Figure 1a, solid circles), with the most mafic sample having an FeO content of 11.7 wt.%. For the other horizons, most samples have FeO>11.4 wt.%, with four notably lower values. One of these four samples (from h2) is the most mafic komatiite chill margin we have sampled with an FeO content of 10.96 wt.%. Though it is from h2, it falls on the FeO fractionation curve defined by the h1 samples. Its higher TiO_2 contents however, show that it is related to the h2-h5 komatiites. The other three low-FeO samples (two from h2, one from h3) could be potential parental magmas to the rest of the h2-h5 magmas. Alternatively, their low FeO could be the result of chemical loss during metamorphism. In sum, we conclude that the least fractionated Kk magmas had FeO contents of 10.0 to 11.7 wt.%, with values >11.0% being the most probable. This estimate is in good agreement with previous estimates (11 wt.% FeO and 29 wt.% MgO; Smith *et al.*, 1980).

The most magnesian olivine we have found in the Komati Formation is Fo94.1 in a sample from h4 (Parman *et al.*, 1997). The most magnesian olivine from h1 is Fo93.9 (unpublished data). Average Fo contents of olivine cores are 93.0 to 93.5. Throughout the Komati Formation, the relict olivine grains have been affected by diffusion of Fe (and presumably Mg) during metamorphism (Parman *et al.*, 1997). This process has raised the FeO content of the relict olivine rims; whether it has also affected the olivine cores is an open question. If so, then the Fo contents of the olivines could have been raised (by diffusion of Fe out of the cores). Profiles across original, igneous olivine grains (now preserved only as a few relict olivines among a network of serpentine veins), are consistent with igneous zoning, so we will assume that the core compositions are unaffected.

With melt FeO contents of 11.0 to 11.7 wt.%, Fo93.9 olivine will be in equilibrium with melts with 29-31 wt.% MgO (Figure 2). The h1 chill margins have MgO contents in this range. Four h1 samples have >31 wt.% MgO, and until olivine with higher Fo contents than 93.9 are found, one must conclude that the MgO content of these samples has been increased by serpentinization. The average olivines (Fo93.0 to 93.5) would be in equilibrium with melts with 25 to 28wt.% MgO, encompassing the majority of the komatiite chill margin data. We consider these to be the most likely komatiite compositions. While those with >28wt.% MgO are possible, more study on the effect of metamorphism on the olivine compositions and on the FeO contents of the bulk samples is required.

We note that the equilibrium curves (Figure 2) are nearly horizontal at high MgO contents, so that small changes in K_D, melt FeO or olivine Fo content can cause large variations in the equilibrium MgO content. Errors on the estimated melt MgO contents could be several weight percent. For instance, if the initial FeO contents were 10 wt.%, instead of 11 wt.%, then the estimated MgO content would shift from 29 to 26 wt.% (assuming Fo94 olivine).

Equilibrium of komatiites with Kaapvaal mantle

Fe-Mg exchange between olivine and melt can also be used to determine whether the Kk could have produced the degree of depletion recorded by the Kaapvaal mantle xenoliths. Pressure increases Fe-Mg K_D (Ulmer, 1989). Our high-pressure, hydrous experiments on komatiitic melt compositions produced K_D values of 0.342 with a 1-sigma standard deviation of 0.01 (Parman, 2001). These values are in excellent agreement with the predictive expression of Ulmer (1989).

Olivines have a range of forsterite contents in Kaapvaal xenoliths. Fertile, sheared lherzolites have Fo 90-92 olivines, with averages around 90.5. Unsheared, harzburgites have Fo 90 up to Fo 95 olivine, with average values of Fo 93.0 (Danchin *et al.*, 1979). The Kk (Mg# up to 0.830) could be in equilibrium with the average harzburgite xenoliths (Figure 3), and so could have produced the degree of depletion in the bulk of the cratonic mantle. But the komatiites are not mafic enough to be in equilibrium with the most depleted harzburgites with Fo 95 olivine. These mantle residues must have been produced by melts even more mafic than the Kk, and may be represented by the extremely mafic ~3.3 Ga Commondale komatiites that outcrop south of the Barberton greenstone belt (Wilson *et al.*, 2003).

Comparison of boninite mantle residue with Kaapvaal xenoliths

If the cratonic mantle was produced by a boninite-like melting process, then one important question is whether or not boninites are producing craton-like mantle today. Much of the fore-arc crust of the Mariana arc contains boninite magmas (Stern and Bloomer, 1992). Serpentine mud volcanoes have brought xenoliths of the Mariana fore-arc mantle to the surface and have been sampled by ODP leg 125 (Ishii *et al.*, 1992). Presumably, these xenoliths represent the residue from the boninite melting process. The samples are all harzburgites and dunites, consistent with the high degree of melting estimated for boninites. The majority of the samples have olivine with Fo ~92 and spinels with Cr# (=Cr/Cr+Al) between 0.50 and 0.65 (Figure 4). Both of these values indicate that the boninite residue is much more depleted than the residues of MORB or OIB melting. Some of the samples, including both harzburgites and dunites, are characterized by much higher extents of melting, with Fo contents up to 94.1 and Cr#'s up to 0.87 (Figure 4). The extent of depletion recorded in these samples overlaps the majority of xenoliths from the Kaapvaal, with only the highest degrees of depletion in the cratonic mantle (Fo>94.5) not being present.

Like the cratonic mantle, the Mariana fore-arc mantle should be buoyant and highly resistant to the homogenizing effects of mantle convection. It has undergone, and will continue to undergo, severe metasomatism due to volatiles released by the subducting Pacific plate, as evidenced by the existence serpentinite mud volcanoes. The fore-arc mantle is also well below typical oceanic geotherms (Uyeda and Horai, 1982). When continent collision closes the Pacific ocean, the fore-arc mantle likely will be accreted under the Eurasian plate. There it may undergo further metasomatism as new convergent margins subduct oceanic crust under the buoyant continent. At that point, it will be cold, highly depleted and complexly metasomatized, similar in all respects to cratonic mantle. As the boninitic fore-arc crust is on the overriding plate, there is a good chance that some will be obducted onto the continental crust, above its mantle residue, as the komatiites now sit above their residues. Though given the horizontal motions involved, there is likely to be some lateral offset between the boninites and their mantle residues. It is also possible that, like modern ophiolites, some part of the mantle residue could be obducted along with the komatiites. Mantle sections in the Barberton greenstone belt have been reported by de Wit (1987).

We propose that the process elucidated above is essentially how the Kk were formed in the Archaean along with the Kaapvaal lithospheric mantle. The proposed process is largely the same as that elucidated by Kesson and Ringwood (1989a; b), Abbott (1991), Jordan (1981; 1988) and de Wit (1998). The main contribution of this article is to propose that it is not normal sub-arc mantle that produces the cratonic mantle, nor is it the residue of hot komatiite-producing mid-ocean ridge magmatism, but is closely linked to transient, subduction-related boninitic magmatism. Due to higher mantle temperatures in the Archaean, ancient boninitic magmatism produced komatiites in addition to boninites (*i.e.* basaltic komatiites) and low-Ti tholeiites. Boninite magmatism has apparently occurred throughout much of Earth history (*e.g.* Kerrich *et al.*, 1998), so one may also speculate that craton-like mantle has also been continuously produced. The depletion produced by Phanerozoic boninites, while close to cratonic mantle levels, may not be quite severe enough to produce the chemical buoyancy required to shield it from mantle convection. Perhaps only komatiite production is sufficient. Or perhaps the rates of boninitic magmatism have severely declined over Earth history. It has been speculated that the length of mid-ocean ridges was much greater in the Archaean, in order to transfer the greater amount of heat in the Archaean mantle. If so, the length of subduction zones would also have been longer. Perhaps much of the world resembled the southwest Pacific with its multiple and complex convergent margins and high rates of arc magma production. Interestingly, and perhaps not coincidentally, that area also contains many of the world's recent boninite occurrences (Crawford *et al.*, 1989).

Pressure-temperature history of cratonic mantle

The above model is admittedly speculative. Yet it agrees well, not only with the geochemisty of komatiites and xenoliths, but also with estimates of the PT history of the Kaapvaal mantle. While the current craton is relatively cold and extends to great depth (Jordan, 1981), the xenoliths record an earlier, hotter history. Garnet (gt) and clinopyroxene (cpx) grains are spatially associated with orthopyroxene in many of the Kaapvaal xenoliths, leading to the hypothesis that the gt and cpx exsolved from the opx as it cooled and was taken to greater pressure (Cox *et al.*, 1987). By adding the gt and cpx back into the opx, the original opx compositions can be estimated and used to constrain the initial PT conditions. Current estimates indicate that the Kaapvaal xenoliths experienced temperatures of 1350 to 1400°C at pressures of 1.5 to 2.5 GPa (Saltzer *et al.*, 2001) at some early stage of their history (Figure 5). These conditions are quite close to the hydrous melt generation conditions for boninite and komatiite magmas (Crawford *et al.*, 1989; Grove *et al.*, 1999; Parman, 2001). Therefore the reconstructed PT history of the xenoliths is consistent with the hypothesis that they were the residues of hydrous melting at shallow levels in a subduction zone, and then were subsequently cooled and taken to greater depths, possibly during collisional tectonism. It is interesting to note that garnet has been implicated in the source of komatiites to explain REE abundance patterns and high CaO/Al_2O_3 ratios (*e.g.* Walter, 1998). Hydrous melting can occur while the mantle is cooling, and so it is possible that the exsolution of garnet from the orthopyroxenes could have occurred while the komatiite magmas were being produced.

Discussion and unanswered questions

While the proposed hypothesis explains many features of the cratonic mantle, there are many that are left unexplained.

(1) Is there enough komatiite in the Kaapvaal crust to balance the depleted cratonic mantle? The answer is probably no. To produce the degree of depletion seen in the craton would require 30-40% melting of a fertile mantle source. If the mantle was already depleted (at a mid-ocean ridge for example), the melting percent could be reduced to perhaps as low as 10%. The depleted cratonic lithosphere is ~150km thick. So to balance this requires a layer of corresponding melt (*i.e.* komatiite) at least 15km thick over the whole craton. The exposed volume of komatiites on the Kaapvaal craton is much less than this and there is no geophysical evidence for such a thick layer of komatiite residing within the Kaapvaal crust. One must conclude that a large proportion of komatiite sequences have been eroded or were never incorporated onto the stable craton.

(2) Why is there no mantle section preserved with komatiites? Many boninites occur in ophiolites, where they are accompanied by sheeted dykes and a peridotite mantle section. Neither of these has been found in association with any komatiites, including the Kk (though see de Wit *et al.*, 1987). Perhaps the thermal gradients in the Archaean crust were such that only the upper portions of the crust were brittle enough to have been thrust onto the continents.

(3) Why are the komatiite sections so thick? While the volcanic sections of boninite-containing ophiolites are ~3km thick, the combined Komati and Hooggenoeg Formations (assuming they are a continuous sequence) is 6km thick. The komatiitic units are restricted to the bottom 3km. So a potential answer is that the higher extents of melting, as evidenced by the komatiites, produced 3km of additional melts. Alternativley, the Komati and Hooggenoeg Formations could be unrelated, as there are numerous shear zones in the section.

(4) What caused the opx-enrichment seen in many xenoliths? Many ideas have been forwarded to explain the high opx contents of xenoliths with Fo rich olivine: melting at high pressures (Kinzler and Grove, 1999), addition of SiO_2 by subduction fluid (Kesson and Ringwood, 1989b) and melt-wallrock reaction (Kelemen *et al.*, 1992; 1998). All of these are possible within the subduction process we have outlined above. We note that many of the boninitic/komatiitic basalt compositions are extremely opx rich. This is due to the harzburgite melting reaction being dominated by opx. If these melts ponded or were trapped in the mantle to varying degrees as they ascended through the lithosphere, they would crystallize a small amount of ol and a large amount of opx, and might produce the opx enrichment seen in the xenoliths without substantially raising (or lowering) olivine Fo contents.

(5) Is the mantle beneath the komatiites of the Kaapvaal similar to the mantle sampled by the Kaapvaal xenoliths? The kimberlites that exhume the mantle samples are in the western part of the craton, hundreds of kilometers from the komatiite outcrops in the eastern Kaapvaal. Seismic tomography indicates that depleted lithospheric mantle underlies both the kimberlites and the komatiites (James *et al.*, 2001). So, like previous researchers, we have assumed that the xenoliths are good proxies for the composition of the eastern Kaapvaal lithospheric mantle, but this is not necessarily the case. Does the mantle residue of the exposed komatiites lie directly beneath them, or have tectonic forces decoupled the crust from the mantle?

(6) Finally, to what extent are greenstone belts like Barberton the Archaean equivalents of Phanerozoic ophiolites? The compositional similarities between boninites and komatiitic magmas strengthens the arguments for a connection between ophiolites and greenstone belts, but the differences (*i.e.* presence of tonalites/trondhjemites, presence of komatiites, thickness of volcanic section, lack of sheeted dykes,...) call this into question. It has been proposed that the Komati and Hooggenoeg Formations are part of an

ophiolite suite called the Jamestown Ophiolite Complex (de Wit *et al.*, 1997) and that ophiolites are present in a number of Archaean greenstone belts (de Wit *et al.*, 2003).

The above questions remain open and point the direction for future studies.

Conclusions

We propose a process by which the generation of komatiites in Archaean subduction zones produced the highly depleted portions Kaapvaal cratonic mantle. The initial melting event was similar to the melting event that produced boninites in the Mariana fore-arc (step 1, Figures 4 and 6). The depleted mantle residue resided in the fore-arc lithosphere where it was cooled and metasomatised over the lifespan of the arc (step 2, Figures 4 and 6). At the end of subduction, presumably by collisional tectonics (step 3, Figures 4 and 6) the fore-arc mantle was thickened by compression and emplaced underneath the overriding plate, similar to the process envisioned by Jordan (1988). The depletion of the komatiite residue and its low temperature (due to cooling by the subducting lithosphere) prevented it from being swept into the induced mantle wedge corner flow of the subduction zone.

The proposed model is consistent with: 1) the evidence for the Kk having been produced in a subduction zone by hydrous melting (Parman *et al.*, 1997), 2) the geochemical similarities between the Kk and modern boninites (Parman *et al.*, 2001), 3) the comparable degree of depletion seen in the Kaapvaal lithospheric mantle and Mariana fore-arc (Figure 4), 4) the estimated PT history of the Kaapvaal xenoliths from reconstructed opx compositions (Saltzer, 2001), 5) the evidence that eclogitic xenoliths were formed from subducted oceanic crust (Schulze, 1986; Shirey *et al.*, 2001), and 6) the overall geology of the Barberton greenstone belt, which indicates a process of crustal growth through arc accretionary processes (de Ronde and de Wit, 1994).

The Izu-Bonin-Mariana (IBM) subduction system is ~2000km long with an estimated width of boninitic magmatism along that length of ~200 km (Stern and Bloomer, 1992). Assuming a depth of melting of 30km for the boninites, one estimates ~$1 \times 10^7 km^3$ of depleted, harzburgitic mantle residue. The Kaapvaal craton has an areal extent of ~$1.2 \times 10^6 km^2$ (de Wit *et al.*, 1998). Assuming that the granular, depleted xenoliths come from a layer that is ~100km thick, the volume of the depleted cratonic mantle is ~$1.2 \times 10^8 km^3$. So it would take approximately 12 IBM sized convergent margins to produce the Kaapvaal lithospheric mantle. The depth of melting for komatiites is estimated to be greater than for boninites, perhaps as deep as 60 to 90 km (Grove *et al.*, 1999; Parman, 2001). If so, then only 4 to 6 arcs would be required to build the Kaapvaal craton. The craton comprises up to 8 individual crustal blocks that amalgamated between ~3.2 and ~2.7 Ga (de Wit *et al.*, 1992). Komatiites are exposed on most of these blocks. However, the purpose of this exercise is not to quantify precisely how many arcs produced the Kaapvaal cratonic mantle. Rather it is to demonstrate that, given the number of exposed greenstone belts with komatiites (though not their preserved volume), it seems plausible that the current volume of the Kaapvaal lithospheric mantle could have grown and been stabilized by the proposed subduction mechanism within mid-Archaean times.

Acknowledgements

We would like to thank all of the researchers involved in the Kaapvaal Project for their encouragement, stimulating discussions and great meetings. The paper was greatly improved by the constructive reviews of Alan Wilson and Hubert Staudigel. This research was supported by National Science Foundation grant EAR-9526702. A. Wilson and H. Standegal critically reviewed an earlier version of this manuscript.

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